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4.6 TECHNIQUES FOR ASSESSING SOIL WATER STATUS IN RAINFED AGRICULTURE

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SUMMARY

Methods are described for measuring soil water content gravimetrically or with the neutron probe, and soil water potential with tensiometers and thermocouple psychrometers. The limitations of each technique are discussed and details are given of their operations in studies of crop water use. Information from tensiometers and the neutron probe can be used together to reliably separate drainage from evaporative losses and to calculate the water transpired by crops. Extraction of water by wheat and millet crops is examined in relation to their pattern of rooting and some of the factors controlling water uptake are examined.

1. INTRODUCTION

Crop production, in many areas of the world, is limited by the lack of water to sustain plant growth. Even in temperate, maritime climates, crops frequently experience short periods of water shortage and yields are reduced (Gallagher, Biscoe and Hunter, 1976; Monteith, 1977). In areas relying on moisture stored in the soil to support a crop, the likelihood of water shortage is acute and a variety of management practices have evolved to make optimum use of water reserves (Janssen, 1972; Spratt and Chowdhury, 1978).

Vital to an improvement of yields is a clearer understanding of the times when water shortage may most affect yield and of the quantities of water used by crops during growth. The effects of water deficits on physiological processes within crops have been reviewed recently by Boyer and McPherson (1975) and Begg and Turner (1976)and the production of crops in arid zones is receiving increasing attention (Fischer and Turner, 1978). Our improved understanding of crop responses to water stress has been accompanied by an improvement in the techniques that enable us to measure soil water status, particularly since the advent of the neutron probe and of thermocouple psychrometry.

The purpose of this paper is to review the methods currently available to measure soil water content and potential, and to indicate the possible difficulties in the interpretation of the measurements. As a specific example, water extraction by wheat and millet is compared to their rooting patterns and the factors controlling water uptake by crops are examined.

2. MEASUREMENT OF SOIL WATER CONTENT

Until about 20 years ago, almost all measurements of soil water content were by direct weighing. A sample of soil at field moisture content is collected and weighed (m_{s+w}), and then dried in an oven at 105°C to constant weight (18 - 24 hour drying). The sample is reweighed (m_s) and the water content on a dry mass basis (θ_m) is calculated from:

$$\theta_m = \frac{m_s + w - m_s}{m_s} = \frac{m_s + w}{m_s} - 1$$

Usually, the water content is expressed as a percentage of the dry mass of soil (P_m), where $P_m = 100 \ \theta_m$.

This method is cheap and simple, although care must be taken to avoid evaporation loss during transit from the field to the laboratory. After sampling, the soil should be placed immediately into sealed metal containers so that any vapour condensing on the inside of the canister can be included in the determination. If samples are collected in polythene bags, waxed paper cartons or other containers which cannot be heated to 105°C without risk or damage, then significant errors will be introduced into the calculation of θ_m. The development of microwave ovens (Miller, Smith and Biggar, 1974) considerably reduces the time taken to dry a sample of soil and this technique may allow the use of inexpensive paper containers (Hankin and Sawhney, 1978).

The disadvantages of gravimetric determination are, firstly, that it is destructive, secondly, that much replication is necessary if values representative of large areas are required; and thirdly, that it is often difficult to obtain samples from soil layers other than the topmost. Moreover, when assessing the use of water by crops or their irrigation requirements, it is more useful to know the water content as a ratio of the volume of water to the volume of soil. This volumetric water content (θ_v) cannot be calculated from the gravimetric water content (θ_m) unless the apparent specific gravity of the soil is known since:

$$\theta_v = \frac{\rho_b}{\rho_w} \ \theta_m$$

where ρ_b is the bulk density of the soil and ρ_w is the density of water.

Volumetric water content can be measured directly using a neutron probe (Gardner and Kirkham, 1952; Long and French, 1967) and this technique has two major advantages: it is non-destructive and it achieves a measure of spatial integration. Aluminium access tubes are installed vertically into the soil and a source of high energy neutrons (usually an americium/berylium source) is lowered into the tube (Fig. 1). Neutrons emitted from the source collide with atomic nuclei in the soil; if these nuclei are large (e.g. aluminium), they bounce off, retaining almost the same velocity. If the nucleii are small, however, the neutrons lose energy to the small nucleii and are slowed down and after several such collisions, the velocity of the neutrons depends on the temperature of the system and such neutrons are said to be "thermalized". The nucleus most effective in thermalizing neutrons is hydrogen which has almost the same mass as a neutron and in soils is usually most abundant as water. Lesser amounts are associated with clay and organic matter. Hence the density of the resultant cloud of slow neutrons around the access tube is a function of the soil water content. The density of the thermalized neutrons is measured by a slow neutron detector (usually a boron trifluoride proportional counter) and the count rate is converted into a volumetric soil water content using an appropriate calibration curve.

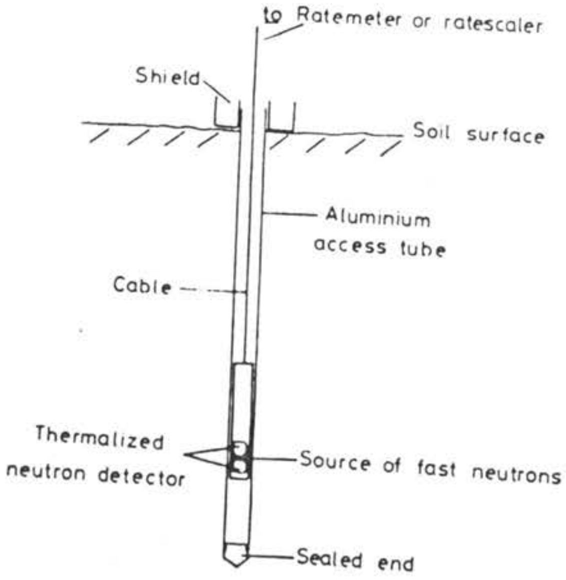


Fig. 1 The neutron probe

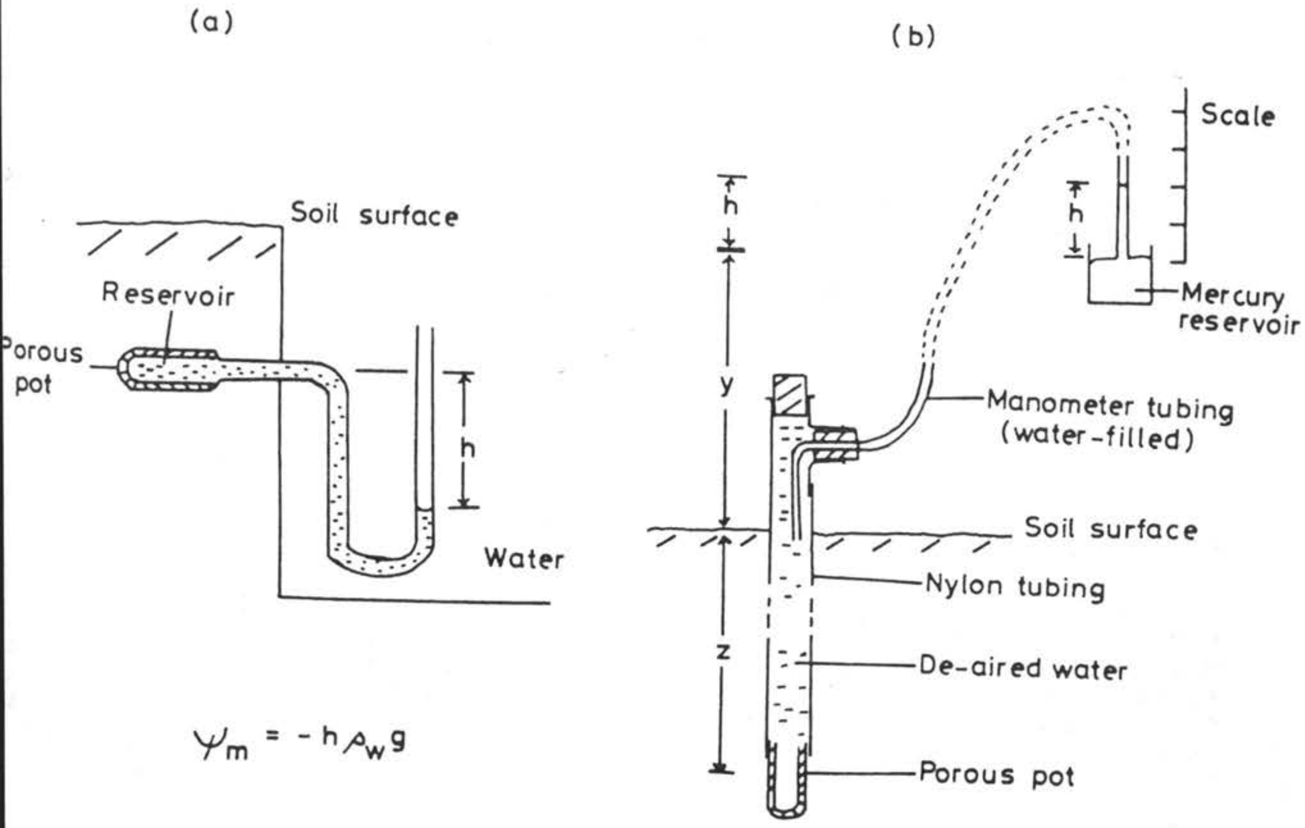


Fig. 2 (a) A simple tensiometer and (b) its typical arrangement for use in the field. The symbols are explained in the text.

Other elements, such as boron, chloride, iron and carbon, may also thermalize neutrons so that the bulk chemical composition affects the count rate. Moreover, the bulk density of the soil may also affect the thermalization of neutrons, so that there is a unique relationship between the water content of a particular soil and the corresponding count rate. The factors affecting the count rate and the problems of calibration have been reviewed by Visvalingam and Tandy (1972) and Bell (1973). Fortunately, in most studies of soil water, one is not interested in absolute values of soil water content, but in changes of water content during an interval of time. In these circumstances, calibration is straightforward because, for most soils, the difference in gradient between calibration lines is small. Fry (1975), studying the water use by crops on a sandy-loam and a clay soil calculated that, if he had used the clay calibration for the sandy soil and vice-versa, the seasonal water balance would have been in error by only 3%.

One of the most serious limitations of the method is that, as the neutron source approaches the soil surface, there is a significant loss of both fast and thermalized neutrons from the soil system. Reliable measurements in the top 20 cm (wet soil) to 30 cm (dry soil) are, therefore, difficult to obtain without modifying the method. The problems can be overcome by: (i) applying correlation factors to the calibration curve; (ii) introducing a reflector at the soil surface; and (iii) artificially raising the surface using an extension tray.

The merits of these alternatives are discussed by Visvalingam and Tandy (1972) and Bell (1973), but at the University of Nottingham, we have successfully adopted the approach of using the probe shield to reduce neutron loss at the soil surface and modifying the calibration curve so that readings are possible at probe depths of 10 cm.

4. MEASUREMENT OF SOIL WATER POTENTIAL

By definition, the total potential of water (Ψ) is the work required to transfer unit quantity of water from a standard reference state (where $\Psi = 0$) to the position where the potential has the defined value. Potential, then, gives an indication of the energy status and hence availability of water since the lower potential, the lower is the availability of water. The total potential (Ψ) is made up of a number of components:

$$\Psi = \Psi_m + \Psi_q + \Psi_o$$

where Ψ_m is the matric potential, Ψ_q is the gravitational potential and Ψ_o is the osmotic potential, and may be specified in terms of hydrostatic pressure or specific free energy (see Appendix).

Matric potential (Ψ_m) can be measured directly by a tensiometer which consists essentially of a reservoir of water enclosed within a porous membrane (normally a ceramic pot) sited within the soil matrix and connected at the other end to either a bourden gauge or a manometer (Fig. 2a). Continuous liquid contact between the soil matrix and tensiometer means that, as the soil dries, water from the reservoir is drawn across the membrane creating a suction that can be measured. In the form shown in Fig. 2a, the tensiometer measures Ψ_m directly, but this necessitates the digging of pits and is, therefore, unsuitable for general field use.

Webster (1966) describes the construction of tensiometer units for use in the field (Fig. 2b). In this arrangement, the reading h on the manometer is

the sum of the matric potential (ψ_m) and the gravitational potential (ψ_g), referred to as the hydraulic potential (ψ_h). Since the height of the mercury reservoir above the ground surface (distance y) varies between workers, readings should be referenced to a stated datum, conveniently the soil surface.
Balancing heads:

$$h + y + z + \psi_m = h \times \rho_{Hg}$$

where ρ_{Hg} is the density of mercury and ψ_m is the matric potential (cm of water). Let h' be the reading when $\psi_m = 0$ and $z = 0$ (i.e. for a water table at the soil surface) then:

$$h' = \frac{y}{12.6}$$

However, when fine-bore tubing is used in the construction of the manometer (to approach the ideal of a null point technique), a correction has to be made for surface tension effects. Using tubing with an internal diameter of about 0.8 mm, mercury rises about 1 cm less than expected.

Thus: $$\frac{y}{12.6} = h' + 1$$

Let y = 25 cm (i.e. an almost exact and convenient multiple of 12.6)

Then h' = 1

Therefore, when quoting the hydraulic potential (cm water) referenced to the soil surface, the appropriate relation is:

$$\psi_h = 12.6\ (h - 1)$$

Temperature will have a marked effect on the operation of the instrument, particularly if there is a temperature difference between the instrument and the soil. The effects of diurnal variations in temperature can be minimized, either by reading the instrument early in the morning before the sun has warmed it to a temperature above that of the soil, or by placing a reflecting screen around the above-ground parts of the instrument.

Tensiometers are cheap and easy to maintain, but have a limited range (0 to -0.08 Mpa) because air eventually enters the ceramic pot, breaking the hydraulic continuity between soil and manometer and the instrument fails. Fortunately, the limited range is the one of most interest when determining irrigation requirements.

The development of thermocouple psychrometers (Monteith and Owen, 1958; Rawlins and Dalton, 1967) has allowed the range of soil water potentials measurable in the field to be extended down to -5.0 MPa with an upper limit of about -0.1 MPa. The unit commonly used consists of a small thermocouple (usually chromel/constantan) enclosed within a ceramic container (Fig. 3a) buried in the soil and is used to measure the vapour pressure in equilibrium with soil. The vapour pressure is a function of soil water and salt contents, and the potential measured with the thermocouple psychrometer (the water potential ψ_w) is the sum of the matric (ψ_m) and osmotic (ψ_o) potentials. Measurement can be made of either the dew point or the wet bulb depression, but in our experience, the latter method is more suitable for field use.

When using the wet bulb depression technique, the thermocouple is first cooled below the dew point by passing a small current through it (Peltier effect) so that a small drop of water condenses on the cooled junction. When the current is discontinued, water evaporates from the wet junction cooling it below the temperature of the reference junction, thereby producing a small e.m.f. The degree of cooling is controlled by the evaporation rate which is, in turn, related to the vapour pressure of the chamber and ultimately the water potential of the soil.

Water potential (Ψ_w) is related to the measured relative humidity (e/e_o) by:

$$\Psi_w = \frac{RT}{V_m} \ln \frac{e}{e_o}$$

where R is the universal gas contant, T is the absolute temperature and V_m is the molar volume of pure water. Table 1 shows that the equilibrium vapour pressure of water at 25°C varies between 0.999 and 0.986 in the range -0.1 to -2.0 MPa and that the corresponding temperature differences between the wet and reference junctions are in the range 0.009 to 0.180°C. The method is obviously very sensitive to fluctuations in ambient temperature and Rawlins and Dalton (1967) have detailed the likely consequences of temperature on psychrometer readings. The main effect of temperature is, if the temperature of the chamber, where the humidity is being measured, changes from that of the thermocouple reference junction measured before the cooling current was passed. At 25°C, this source of error amounts to about 11 MPa per °C so that temperature differences of less than 0.001°C are necessary to measure soil water potentials with an accuracy of 0.01 MPa. Thus, when installing thermocouple psychrometer units in the field, it is advisable to minimize temperature gradients within the system, particularly in the surface layers of soil where large diurnal fluctuations in temperature occur. Valancogne and Daudet (1974) showed the effect of positioning on the stability of psychrometer units and concluded from measurements and theoretical calculations, that units, like those shown in Fig. 3a, should be installed horizontally in the soil.

Table 1 THE RELATIONSHIP (AT 25°C) BETWEEN WATER POTENTIAL, VAPOUR PRESSURE AND WET BULB DEPRESSION, AND THE CONSEQUENCES OF AN ERROR IN THE MEASUREMENT OF WET BULB DEPRESSION ON THE ESTIMATE OF WATER POTENTIAL

Ψ (MPa)	e/e_o	ΔT (OC)	Ψ if error of 0.01°C in T (MPa)
-0.1	0.99928	0.0090	-0.21
-0.2	0.99855	0.0180	-0.31
-0.5	0.99638	0.0450	-0.61
-1.0	0.99278	0.0900	-1.11
-1.5	0.98920	0.1350	-1.61
-2.0	0.98561	0.1800	-2.11

More recent work of Campbell (1979) has shown that stability can be improved if the reference junction is surrounded by a mass with high thermal conductivity. Such units (Fig, 3b) are less sensitive to axial temperature gradients and can, therefore, be installed vertically. This is a considerable advantage as soil disturbance at the time of installation is then minimal.

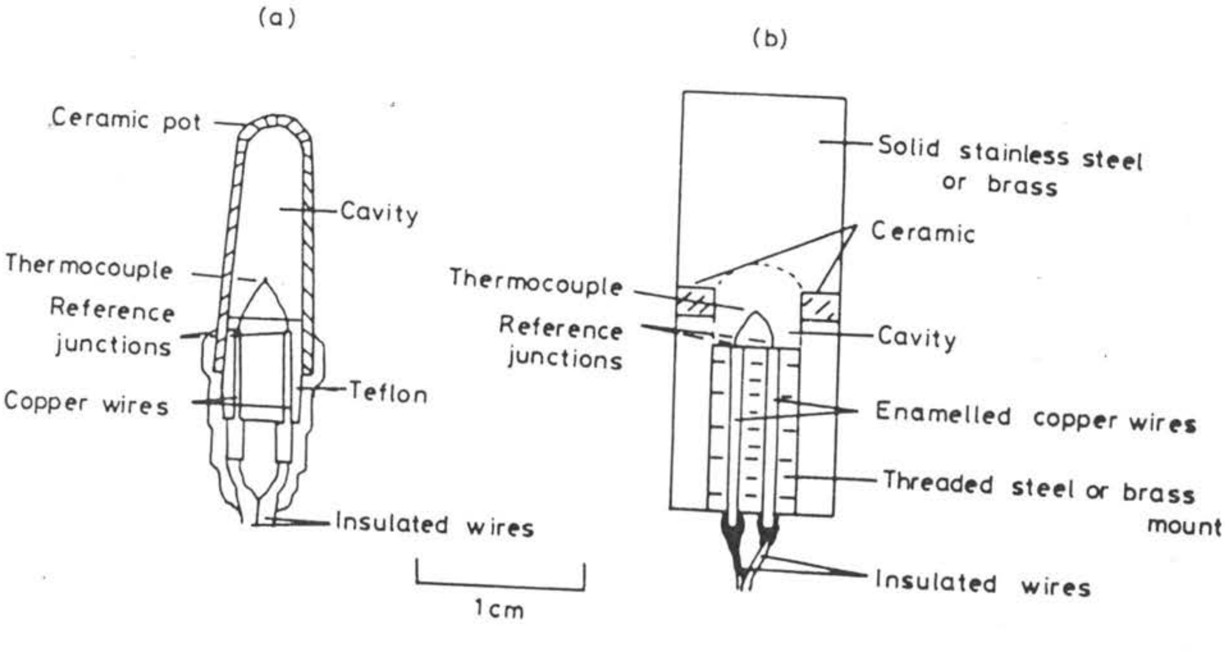


Fig. 3 Thermocouple psychrometers: (a) typical commercial type and (b) Campbell type

The water potential measured with a thermocouple psychrometer is the sum of the matric and osmotic potentials, but in non-saline soils, the osmotic potential component will be small and in a typical English agricultural soil at a water potential of -1.5 MPa, ψ_o was -0.05 MPa (Gregory, McGowan and Biscoe, 1978). When it is desirable to separate matric and osmotic potential components, then osmotic potential can be measured directly using a modified thermocouple psychrometer (Oster, Rawlins and Ingvalson, 1969). Alternatively, it can be estimated from measurement of the electrical conductivity of a saturation extract (E C mmhos/cm) assuming all solutes remain in solution, using:

$$\psi_o = 0.36 \times E.C. \times \frac{\theta_{v\ sat}}{\theta_v}$$

where $\theta_{v\ sat}$ and θ_v are the volumetric water contents of the soil at saturation and at the measured ψ_w, respectively (U.S.D.A., 1954).

A symposium concerned with the use of thermocouple psychrometers (Brown and van Haveren, 1972) gives various applications of the method in soil and plant water research.

5. ESTIMATION OF SOIL WATER BALANCE

While the neutron probe can be used to measure the changes of water stored in given soil layers with time, additional information is required before the use of water by a crop can be calculated. The amount of water transpired by a crop in a specified time is:

$$T = P - R - E_s - \Delta_S - D$$

where P is precipitation, R is the runoff, E_s is the evaporation from the soil surface, Δ_S is the increase in stored soil water and D is the drainage.

For many purposes (for example, in hydrological studies), it is unnecessary to distinguish between T and E_s and the two are summed as evaporation (E).

The precipitation (P) can be easily measured with a raingauge, but runoff (R) is difficult to measure, although it is unlikely to be important, except when the rate of precipitation exceeds the infiltration rate, or in areas where local redistribution of water occurs. The separation of drainage from evaporation, however, can be a formidable problem (van Bavel, Brust and Stirk, 1968), especially when rain showers rewet the topsoil or where there is a receding water table. Both of these circumstances are common in areas of rain-fed agriculture and to overcome this difficulty, a variety of procedures has been adopted. In deep, uniform soils, with no water table, it is possible to calculate drainage (D) or upward flux of water from a knowledge of the hydraulic conductivity (K) of the soil using:

$$D = K \frac{d\psi}{dz}$$

This approach has been used to calculate water uptake of irrigated cotton (Rose and Stern, 1967) and Stone, Horton and Olson (1973a) have shown the importance of estimating the flux below the root zone when determining evaporation rates using depletion methods.

An alternative approach, adopted in our own work (McGowan, 1974; Gregory, McGowan and Biscoe, 1978), has been to define an "effective rooting depth", above which, all water losses are by evaporation, and below which, water losses are by drainage. This approach is more dynamic than that of Stone et al.(1973b) in that the effective rooting depth may change through the growing season, while the alternative method has a fixed boundary. Moreover, it can be used on soils where lateral movement of water occurs or where the composition is very variable. Many soils do not have uniform hydraulic properties and considerable lateral variation can exist in small distances. The principles of the methods have been outlined by McGowan (1974) and the water balance, so calculated, relies on the use of both tensiometer and neutron probe readings.

Figure 4 shows profiles of hydraulic potential referenced from the soil surface for selected days under a crop of winter wheat that had been covered to exclude rain. Since, by definition, water will flow only from zones of high to zones of low potential, the effective rooting depth may be found by identifying the depth of zero hydraulic gradient. On 18 April the potentials at 20,30 cm were lower than at 40 cm, so water above 40 cm must have been moving upwards, i.e. 40 cm represents the effective rooting depth. If tensiometer measurements are made frequently enough, it is possible to define the effective rooting depth accurately throughout the growing season.

Changes of soil water content, measured with the neutron probe under the same crop, are shown in Fig. 5. All layers are losing water throughout the 8-week period shown and it is not immediately apparent whether the losses are by evaporation or drainage. However, as a layer of soil drains, its hydraulic conductivity decreases and, therefore, the rate of water loss from that layer also decreases. When extraction of water by roots commences, the rate of water loss suddenly increases resulting in a discontinuity in the curve of water content plotted against time (Williams, 1971; McGowan, 1974). The identification of these discontinuities provides another method for determining the effective rooting depth and hence a means of separating drainage from evaporation. The effective rooting depths determined from neutron probe and tensiometers are shown in Fig. 6 and indicate comparability. Where discontinuities in the water content against time curves are uncertain, the profiles of hydraulic potential may assist in the interpretation of the measurements.

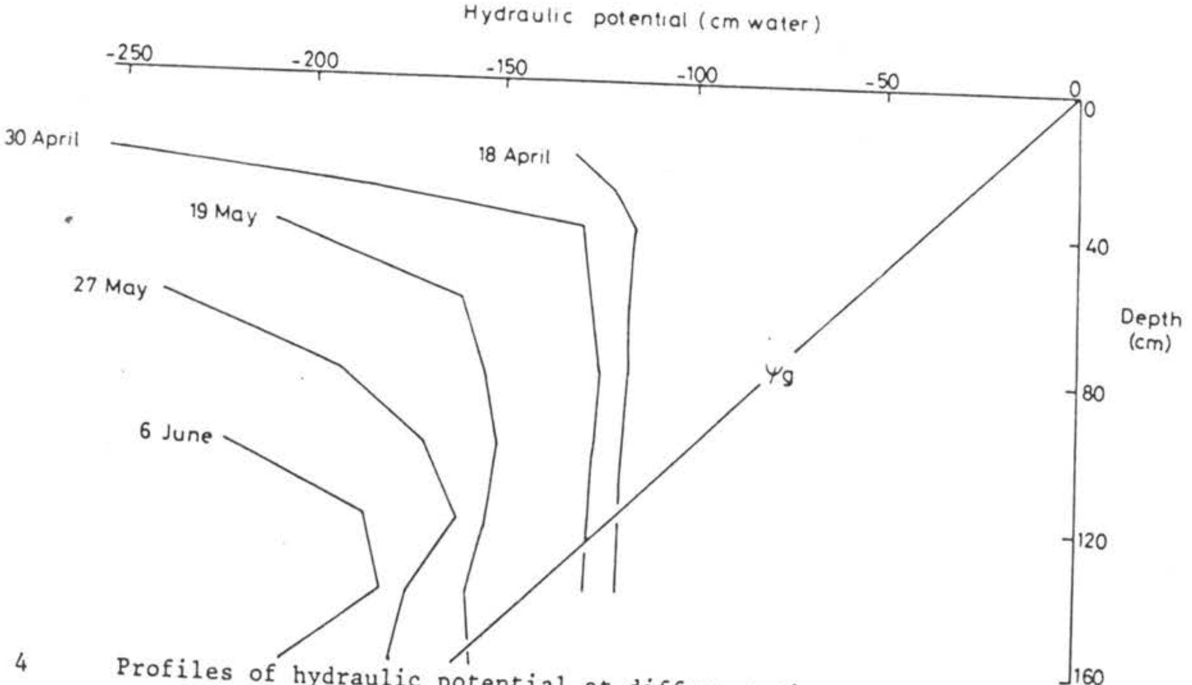
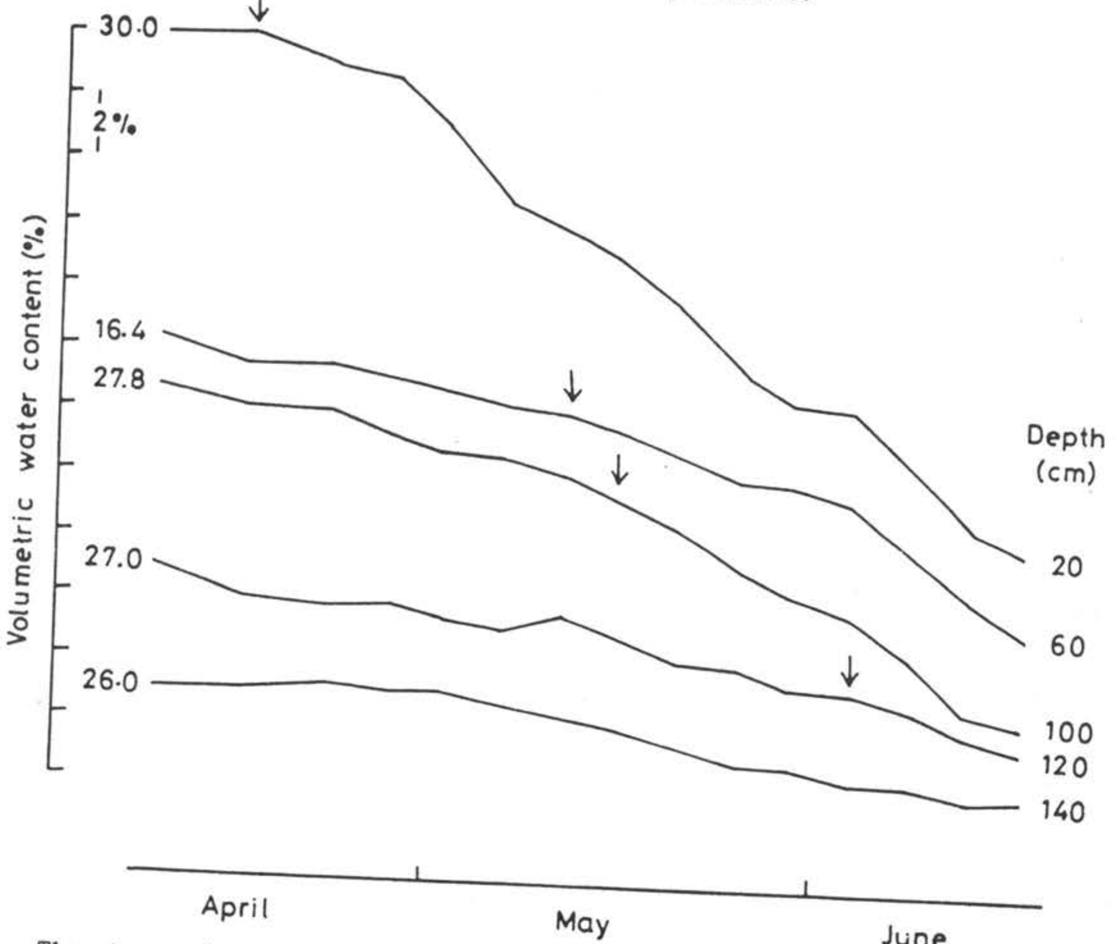


Fig. 4 Profiles of hydraulic potential at different times during growth measured under winter wheat growing on stored soil water. The line ψ_g is the gravitational potential.



ig. 5 The change in soil water content at selected depths measured at different times during growth under winter wheat growing on stored soil water. The initial water content of each layer is shown at the start of each curve and the arrow shows the beginning of water extraction by roots.

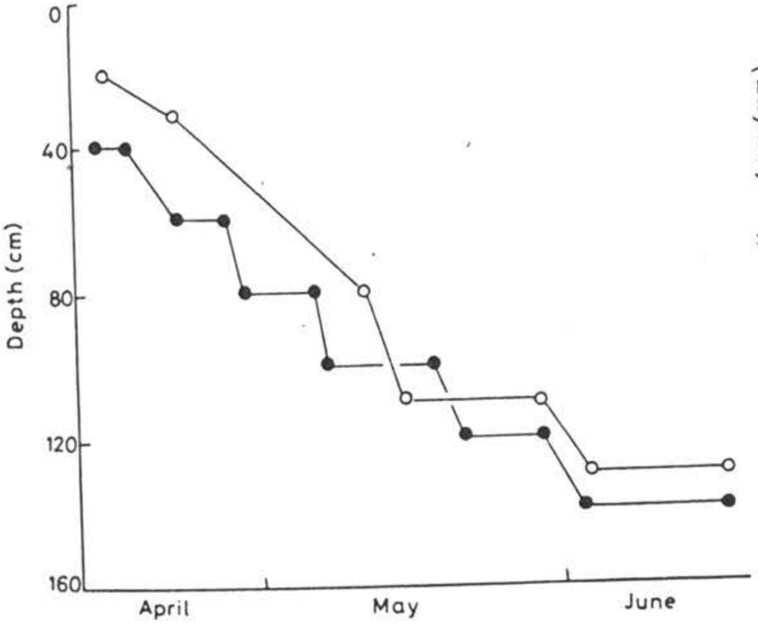


Fig. 6 A comparison of effective rooting depth estimated by tensiometers (●) and the neutron probe (o) for winter wheat growing on stored soil water

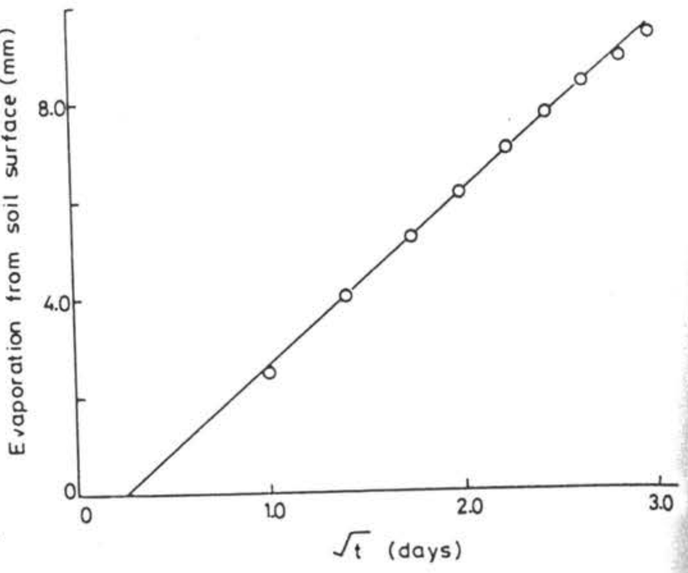


Fig. 7 Cumulative evaporation from bare soil beneath a crop of millet with a LAI of 1.5

In semi-arid regions, E_s may be an important component of E before the plant canopy covers the soil surface, and values given by Fisher and Turner (1978), in four studies, show E_s ranging from 0.2 to 0.5 of E. Kristensen (1974) investigated the dependance of E_s on leaf area index (LAI) and concluded that, for wheat with LAI greater than 4, E_s was less than 10% of E. For low LAI when the surface is wet, E_s is dependent on the amount of radiation reaching the soil surface. As the surface dries, E_s becomes more dependent on the hydraulic conductivity of the surface layers and during a drying cycle:

$$E_s = c\sqrt{t}$$

where c is a constant which must be determined for each soil and may vary seasonally.

E_s is difficult to measure directly, but models for predicting evaporation from row crops with incomplete covers are now available, (Ritchie, 1972) and give good agreement with measurements obtained using lysimeters (Tanner and Jury, 1976). Figure 7 shows the evaporation from the soil surface under a crop of millet with an LAI of about 1.5 measured by weighing large containers, 9 cm deep, packed with soil. The figure shows that, after an initial period, the cumulative loss of water is linearly related to $\sqrt{t}$.

6. WATER USE BY CROPS

A greater appreciation of the factors regulating the water loss from crops has emerged during the last 25 years, particularly since the elucidation of the factors controlling the atmospheric demand for water. Potential evaporation can be calculated from easily measured meteorological variables and used to predict the likely water requirements of crops, or, where water is limited, the severity of drought. Many theoretical and empirical relationships exist to calculate this potential demand (for a brief review, see Taylor and Ashcroft, 1972), but the factors limiting the supply of water in the soil are less well understood.

The concept of "available water" (i.e. water held between -0.03 MPa and -1.5 MPa) gives a useful indication of the amount of water available which roots may be able to extract over the whole of growing season, but has serious limitations when used for shorter-term estimates of periods of water stress. More dynamic concepts of water availability now formulated (Gardner, 1960) show the importance of hydraulic conductivity and the distribution of roots in determining the amounts of water available to plants from hour to hour. However, despite the plethora of water uptake measurements made since the advent of the neutron probe, there are few concurrent measurements of root distributions. This omission reflects the effort required to isolate roots; but systematic root measurements are a prerequisite for a more thorough interpretation of patterns of soil water depletion.

Because root distribution is difficult to measure, several workers have attempted, as a first step, to correlate rooting depth with root activity in extracting water; for short-season crops relying largely on stored moisture, the agreement is generally within 10-15 cm (Durrant *et al.*, 1973; McGowan, 1974; Stone *et al.*, 1976). However, the rate of water uptake from individual soil layers is rarely a good guide to root distribution (Rickman, Allmaras and Ramig, 1978). Figure 8 compares the profiles of root distribution and rates of water uptake for a crop of millet where soil water potentials were maintained between -0.03 and -0.1 MPa until the fourth stage of growth shown. It is clear that, apart from the first period, water uptake and root distribution are unrelated and this is also shown (Table 2) by the crop of winter wheat, mentioned previously, growing into stored moisture. Initially, uptake may

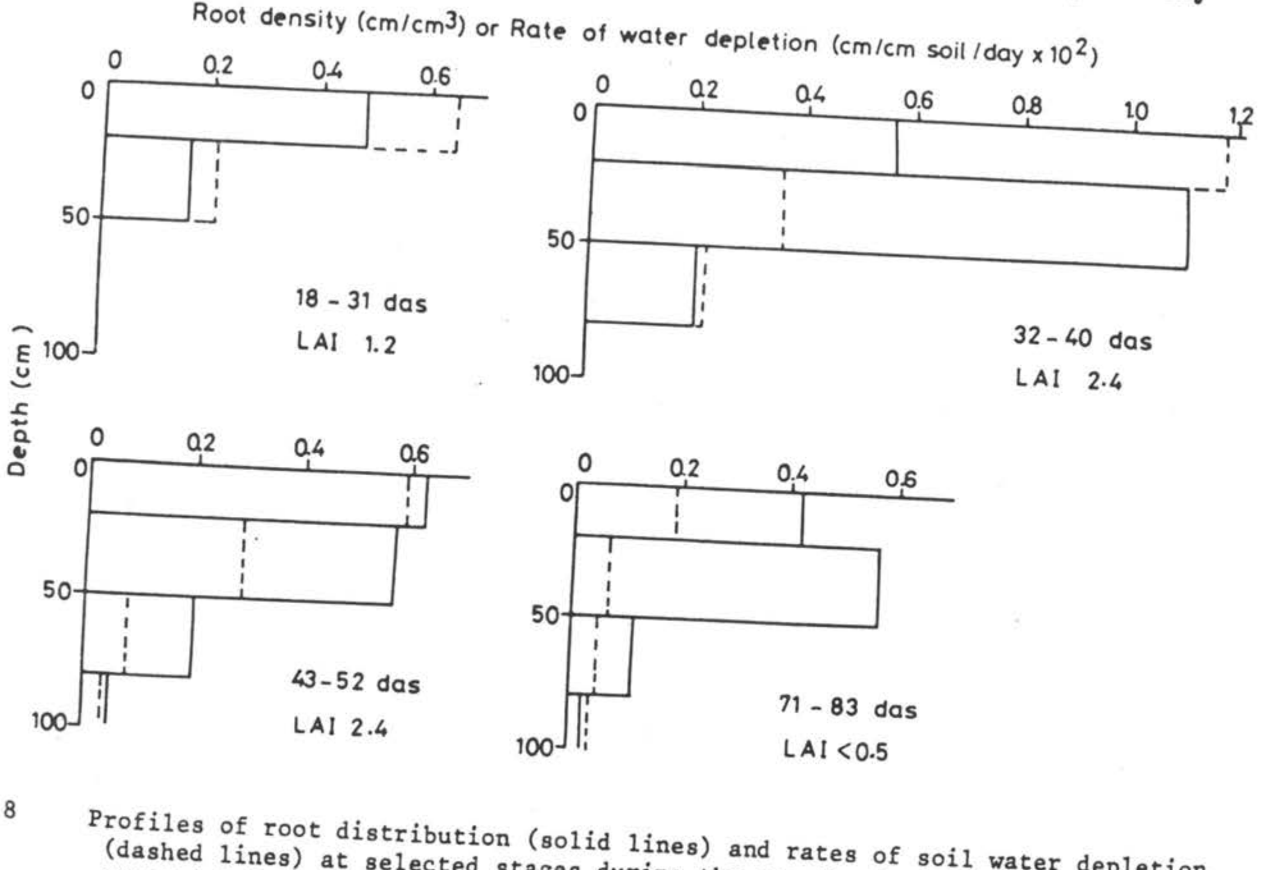


Fig. 8 Profiles of root distribution (solid lines) and rates of soil water depletion (dashed lines) at selected stages during the growth of millet. Soil water potential in the top 0-20 cm was maintained at greater than -0.1 MPa until 52 das by irrigation but was lower than -0.1 thereafter

correspond to distribution, but because of the non-uniform distribution of roots, the layer with most roots will be depleted most rapidly, and the hydraulic conductivity will fall, resulting in the zone of maximum uptake changing throughout growth. Rickman et al. (1978) show the depth of maximum water extraction moving gradually down the soil profile as a crop of wheat grows on stored soil water. Table 2 also shows this phenomenon, as well as the ability of a few, deep roots to supply a substantial proportion of water during a drying phase (see Stone et al., 1976).

Table 2 COMPARISON OF THE RELATIVE WATER USE FROM DIFFERENT SOIL LAYERS WITH THE RELATIVE DISTRIBUTION OF ROOTS AT TWO STAGES IN THE GROWTH OF WINTER WHEAT ON STORED SOIL WATER

		Depth (cm)			
Date		0–30	30–60	60–10	100
20–27 May	root distribution (%)	50	18	25	7
	water uptake (%)	54	16	24	6
10–17 June	root distribution (%)	85	8	6	1
	water uptake (%)	17	30	39	14

In regions where crops are planted into wet soils that dry rapidly because of a high atmospheric demand for water, crop growth will be determined largely by the ability of the plant to approach, rather than actually achieve, the potential rate of evaporation. A key process, enabling this to happen, would be the downward penetration of roots into wet soil, proceeding more rapidly than the downward penetration of the drying front. Many soils may contain appreciable quantities of water below 1 m after a cropping season (Hurd, 1974, wheat, Canada; Russell, pers. comm., millet, India) and varieties are being bred with roots having faster rates of penetration and deeper rooting systems (Hurd, 1974). Since, in many situations, the water use efficiency of crops (i.e. g dry matter produced per g water transpired) grown where water is limiting, appears to be constant (de Wit, 1958), though dependent on crop type (Fischer and Turner, 1978), then a deeper root system with more water accessible should produce higher yields. If, however, the quantity of water in the soil is strictly limited and there are no deep, subsoil reserves, the timing of water availability on relation to yield determining processes will be important (Tennant, 1976). In such circumstances, a sparser root system might allow the limited water to be used more effectively in producing yield (Passioura, 1974).

7. CONCLUSIONS

The development of the neutron probe and thermocouple psychrometer to measure soil water content and potential have allowed more accurate determinationas of the water use by crops and of the gradients of potential in the soil, plant, atmospheric system. Furthermore, the magnitude of water fluxes through plants to the atmosphere is becoming more clearly understood and management strategies are being adopted to take account of these. There remains considerable scope for improving yields, either by breeding to increase the quantity of water available to the root system, or alternatively, by using limited amounts of water more effectively.

Note on Units

Many scientific bodies have now adopted the S.I. unit of pressure, the pascal (Pa) in preference to the c.g.s unit of the bar, which is commonly used in the literature of soil and plant water relations.

Since

$$1 \text{ bar} = 10^5 \text{ Pa}$$

it is convenient to adopt the megapascal

$$1 \text{ MPa} = 10 \text{ bar}$$

as an S.I. unit of appropriate size.

The chemical potential of water used in the thermodynamic analysis of water movement is specified by specific free energy and

$$1 \text{ MPa} = 1 \text{ MJ/m}^3 = 1\text{J/cm}^3$$

Between 0 and 44°C, the density of water lies between 1.000 and 0.990 g/cm^3, so with an accuracy of 1%

$$1 \text{ MPa} = 1 \text{ J}/g$$

When referring to the hydraulic potential registered by tensiometers, it is convenient to quote it in units of cm of water, where:

$$1020 \text{ cm water} = 1 \text{ bar} = 0.1 \text{ MPa}$$

Occasionally, in the literature, the potential of soil water is given as a suction in units of pF, where:

$$\text{pF} = -\log(\text{cm water pressure})$$

i.e. pF3 = - 1000 cm water pressure = - 0.1 MPa.

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REFERENCES

Van Bavel, C.H.M., Brust, K.J. and Stirk, G.B. (1968). Hydraulic
1968 properties of a clay loam soil and the field measurement of water uptake by roots: II. The water balance of the root zone. Soil Sci., Soc, Amer. Proc. 32, 317-321.

Begg, J.E. and Turner, N.C. Crop water deficits. Adv. Agron. 28, 161-217.
1976

Bell, J.P. Neutron probe practice. N.E.R.C. Institute of Hydrology,
1973 Report No. 19.

Boyer, J.S. and McPherson, H.G. Physiology of water deficits in cereal crops.
1975 Adv. Agron. 27, 1-23.

Brown, R.W. and Van Haveren, B.P. Psychrometry in water relations research.
1972 Utah State University.

Campbell, G.S. An improved thermocouple psychrometer for measuring water
1979 potential. J. Phys. E. (In press).

Durrant, M.J., Love, B.T.G., Messem, A.B. and Draycott, A.P. Growth of
1973 crop roots in relation to soil moisture extraction. Ann. Appl. Biol. 74, 387-394.

Fischer, R.A. and Turner, N.C. Plant productivity in the arid and semi-arid
1978 zones. Ann. Rev. Plant Physiol. 29, 277-317.

Fry, G.J. The physical damage of clay soils and the response of
1975 roots. Ph.D. Thesis, University of Nottingham.

Gallagher, J.N., Biscoe, P.V. and Hunter, B. Effects of drought on grain
1976 growth. Nature, Lond. 264, 541-542.

Gardner, W.R. Dynamic aspects of water availability to plants.
1960 Soil Sci, 89, 63-73.

Gardner, W.R. and Kirkham, D. Determination of soil moisture by neutron
1952 scattering. Soil Sci. 73, 391-401.

Gregory, P.J., McGowan, M. and Biscoe, P.V. Water relations of winter
1978 wheat 2 Soil water relations. J. agric. Sci. Camb. 91, 103-116.

Hankin, L. and Sawhney, B.L. Soil moisture determination using microwave
1978 radiation. Soil Science, 126, 313-315.

Hurd, E.A. Phenotype and drought tolerance in wheat. Agric. Met. 14,
1974 39-55.

Janssen, B.H. The significance of the fallow year in the dry farming
1972 system of the Great Kenya Basin, Turkey. Neth. J. Agric. Sci. 20, 247-260.

Kristensen, K.J. Actual evapotranspiration in relation to leaf area.
1974 Nordic Hydrology 5, 173-182.

Long, I.F. and French, B.K. Measurement of soil moisture in the field by
1967 neutron moderation. J. Soil Sci. 18, 149-166.

McGowan, M. Depths of water extraction by roots. Application to soil-
1974 water balance studies. In Isotope and Radiation Techniques in Soil Physics and Irrigation Studies 1973. pp. 435-445. Vienna: I.A.E.A.

Miller, R.J., Smith, R.B. and Biggar, J.W. Soil water content: microwave
1974 oven method. Soil Sci. Soc. Amer. Proc. 38, 535-537.

Monteith, J.L. Climate and the efficiency of crop production in Britain.
1977 Phil. Trans. R. Soc. Lond. B 281, 277-294.

Monteith, J.L. and Owen, P.C. A thermocouple method for measuring relative
1958 humidity in the range 95 - 100%. J. Sci. Instrum. 35, 443-446.

Oster, J.D., Rawlings, S.L. and Ingvalson, R.D. Independent measurement of
1969 matric and osmotic potential of soil water. Soil Sci. Soc. Amer. Proc. 33, 188-192.

Passioura, J.B. The effect of root geometry on the water relations of
1974 temperate cereals (wheat, barley, oats). In Structure and Function of Primary Root Tissue ed. J. Kolek. pp. 357-363. Slovak. Acad. Sci. Czeckoslovakia.

Rawlins, S.L. and Dalton, F.N. Psychrometric measurement of soil water
1967 potential without precise temperature control. Soil Sci. Soc. Amer. Proc. 31, 297-301.

Rickman, R.W., Allimaras, R.R. and Ramig, R.E. Root-sink description of water
1978 supply to dryland wheat. Agron. J. 70, 723-728.

Ritchie, J.T. Model for predicting evaporation from a row crop with
1972 incomplete cover. Water Resources Res. 8, 1204-1213.

Rose, C.W. and Stern, W.R. Determination of withdrawal of water from soil by
1967 crop roots as a function of depth and time. Aust. J. Soil Res. 5, 11-19.

Russell, M.B. personal communication. Soil physicist, ICRISAT, Hyderabad,
1978 India.

Spratt, E.D. and Chowdbury, S.L. Improved cropping systems for rainfed
1978 agriculture in India. Field Crops Research 1, 103-126.

Stone, L.R., Horton, M.L. and Olsen, T.C. Water loss from an irrigated sorghum
1973a field. I. Water flux within and below the root zone. Agron. J. 65, 492-495.

Stone, L.R., Horton, M.L. and Olsen, T.C. Water loss from an irrigated sorghum
1973b field. II. Evapotranspiration and root extraction. Agron. J. 65, 495-497.

Stone, L.R., Teare, I.D., Nickell, C.D. and Mayaki, W.C. Soyabean root
1976 development and soil water depletion. Agron. J. 68, 677-680.

Tanner, C.B. and Jury, W.A. 1976 Estimating evaporation and transpiration from a row crop during incomplete cover. Agron. J. 68, 239-243.

Taylor, S.A. and Ashcroft, G.L. 1972 In physical edaphology. The physics of irrigated and non-irrigated soils. W.H. Freeman and Co., San Francisco pp. 45-84.

Tennant, D. 1976 Wheat root penetration and total available water on a range of soil types. Aust. J. Exp. Agric. An. Husb. 16, 570-577.

United States Department of Agriculture. 1954 Saline and alkali soils. U.S.D.A. Handbook No. 60.

Valancogne, C. and Daudet, F.A. 1974 Adaptation de la technique micropsychrométique à l'enregistrement du potentiel hydrique dans le sol en place: problèmes posés par la température. Ann. agron. 25, 733-751.

Visvalingam, M. and Tandy, J.D. 1972 The neutron method for measuring soil moisture content - A review. J. Soil Sci. 23, 499-511.

Webster, R. 1966 The measurement of soil water tension in the field. New Phytol. 65, 249-258.

Williams, J.B. 1971 The water balance of an agricultural catchment. Ph.D. Thesis, University of Nottingham.

de Wit, C.T. 1958 Transpiration and crop yields. Versl. Landbouwk. Onderz. 64.6, 1-88.